ENERGY EXPENDITURE AND GEOMORPHIC WORK OF THE CATACLYSMIC MISSOULA FLOODING IN THE COLUMBIA RIVER GORGE, USA

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ABSTRACT

Cataclysmic releases from the glacially dammed Lake Missoula, producing exceptionally large floods, have resulted in significant erosional processes occurring over relatively short time spans. Erosional landforms produced by the cataclysmic Missoula floods appear to follow a temporal sequence in many areas of eastern Washington State. This study has focused on the sequence observed between Celilo and the John Day River, where the erosional features can be physically quantified in terms of stream power and geomorphic work. The step-backwater calculations, in conjunction with the geologic evidence of maximum flow stages, indicate a peak discharge for the largest Missoula flood of $10\times10^6\,m^3\,s^{-1}$. The analysis of local flow hydraulics and its spatial variation were obtained calculating the hydrodynamic variables within the different segments of a cross-section. The nature and patterns of erosional features left by the floods are controlled by the local hydraulic variations. Therefore, the association of local hydraulic parameters with erosional and depositional flood features was critical in understanding landform development and geomorphic processes. The critical stream power required to initiate erosion varied for the different landforms of the erosional sequence, ranging from $500\,W\,m^{-2}$ for the streamlined hills, up to $4500\,W\,m^{-2}$ to initiate processes producing inner channels. Erosion is possible only during catastrophic floods exceeding those thresholds of stream power below which no work is expended in erosion. In fact, despite the multiple outbursts which occurred during the late Pleistocene, only a few of them had the required magnitude to overcome the threshold conditions and accomplish significant geomorphic work.

Earth surf. processes landf., **22,** 457–472 (1997)
No. of figures: 8 No. of tables: 0 No. of refs: 48
KEY WORDS: floods; palaeohydrology; stream power; Missoula flooding

INTRODUCTION

During the late Wisconsin, repeated outbursts from ice-dammed glacial Lake Missoula produced cataclysmic floods catalogued as the largest known terrestrial freshwater flows. Descriptions of erosional and depositional landforms produced by the Missoula floods began in the 1920s when (Bretz 1923, 1924, 1925, 1928) studied the channelled scabland in eastern Washington. During recent decades, however, new efforts have been made towards a quantitative understanding of the flood physics, in both palaeohydrological calculations (Baker, 1973a, 1982; Craig, 1987; O'Connor and Baker, 1992; Benito and O'Connor, 1991) and an understanding of flood processes (Baker, 1973b; Baker and Komar, 1987). Erosional landforms produced by the cataclysmic floods appear to follow a temporal sequence described qualitatively by Baker and Komar (1987). The sequence begins with washing out of the loess capping the Columbia Plateau basalts producing streamlined hills. The next stage of the sequence involves the formation of longitudinal grooves, potholes, butte-and-basin topography and eventually inner channels within the basalt surface. This erosional evolution is similar to observations made in experimental studies in flumes with simulated bedrock fluvial systems (Shepherd and Schumm, 1974). As indicated by Baker and Komar (1987), this qualitative sequence requires more precise quantification relating to the energy expended by the flood in producing erosional and depositional features. Furthermore, the understanding of flood physics in the formation of erosional and depositional landforms is complicated by the multiple outbursts from glacial Lake Missoula proposed by Bretz (1969), Bretz *et al*. (1956), Waitt (1980, 1984, 1985), Baker and Bunker (1985) and Benito and O'Connor (1995). Therefore, a two-fold matter is still unresolved: (1) a more precise quantification of erosional and depositional processes relating the

CCC 0197-9337/97/050457–16 $17.50

preserved flood features with the local hydraulic conditions; and (2) the role of the repeated outbursts in modifying previous flood features.

The purpose of this study is to provide an analysis of local flow hydraulics, and their spatial variations and relationships with erosional and depositional landforms. The study area is located in the Columbia River Gorge where the multiple flood pathways of the channelled scabland converge into a single path. The results of step-backwater calculations allow the preliminary assessment of energy conditions associated with erosion and deposition features. Scabland topography or areas of intense erosion are located at sites with unit stream power of up to 100 000 $W m^{-2}$. The erosional landforms described by Baker (1973a,b) and Baker and Komar (1978) within a temporal sequence are found within the Gorge. Ranges of estimated flow depths, flow velocities, shear stresses and stream powers associated with this sequence are suggested. Although depositional features are ubiquitous within the Gorge, only longitudinal bars can be related to energy conditions with the step-backwater modelling. Along the Columbia River Gorge, longitudinal bars are located in reaches of unit stream power magnitudes lower than 15000 $W m^{-2}$. Between Celilo and the John Day River, calculated local hydraulic parameters associated with gravel and pebble bar deposits indicate flow velocities under 8 $m s^{-1}$ and shear stress and stream power below 200 $N m^{-2}$ and 1500 $W m^{-2}$, respectively.

METHODOLOGY

A systematic mapping of erosional and depositional features was performed using aerial photographs, 1:20 000 in scale. Heights and stages of flood features were described by mapping onto US Geological Survey 7·5 minute topographic maps (scale 1:24 000), which have a contour interval of 6 to 12 m. Field excursions were undertaken to conform map and photo interpretations, to evaluate existing data, and to determine the minimum possible altitude range of flood stages.

The discharge estimation associated with geologic evidence was calculated using US Army Corps of Engineers HEC-2 Water Surface Profile computer program (Feldman, 1981; Hydrologic Engineering Center, 1985). The computational procedure is based on the solution of the one-dimensional energy equation, derived from the Bernoulli equation, for steady gradually varied flow. This procedure accounts for energy expended by the flow between discrete cross-sections. These energy losses are the estimated flow-friction losses associated with channel roughness (Manning's *n*), and the form losses from channel expansions and contractions. The palaeohydrological reconstruction is based on the calculation of a step-backwater profile producing the best correlation with the geological evidence of flow stage. Between Portland and Arlington, 122 cross-sections were measured from US Geological Survey topographic maps. Manning's *n* values of 0·05 over the valley floor and 0·07 over the valley margins were assigned. The sensitivity tests performed with different roughness values indicate that uncertainties are of minimal importance to the modelled discharge results. However, energy losses due to expansion and contraction coefficients are critical to the discharge solution. The transition loss is computed by multiplying these coefficients by the absolute difference in velocity heads between two adjacent cross-sections. The loss from short abrupt transition is larger than that from gradual transition. The best correlation of the step-backwater profile with the geological indicators was obtained for contraction–expansion coefficients of 0·1 and 0·3. For the Bonneville flow, contraction and expansion coefficients of 0·0 and 0·5 were used by Jarret and Malde (1987) and O'Connor (1993).

Despite the limitations of a one-dimensional model, such as the use of a uniform energy slope for the entire cross-section, the step-backwater modelling provides an important approximation of the local flow hydraulics and their spatial variation. Velocity, v ($m s^{-1}$), distribution within a cross-section is calculated on the basis of the local flow depth, D (m), local estimated roughness, n, and average energy slope for the cross-section, S_e. From these values the calculation of the boundary shear stress τ ($N m^{-2}$) or tangential force exerted by the flow is straightforward ($\tau = \gamma D S_e$, where γ is the specific weight of the fluid, here assumed to be equal to that of clear water, 9800 $N m^{-3}$), as well as the stream power per unit area, ω ($W m^{-2}$), or time rate of energy expenditure per unit area ($\omega = \tau v$). Unit stream power has been suggested as a good index of sediment transport and geomorphic effectiveness (Bagnold, 1966; Williams, 1983; Costa, 1983; Baker and Costa, 1987). As indicated by O'Connor (1993), the unit stream power is directly related to the local rates of energy expenditure and reflects the physical capability for performing geomorphic work. The reconstruction of these flow conditions in conjunction with

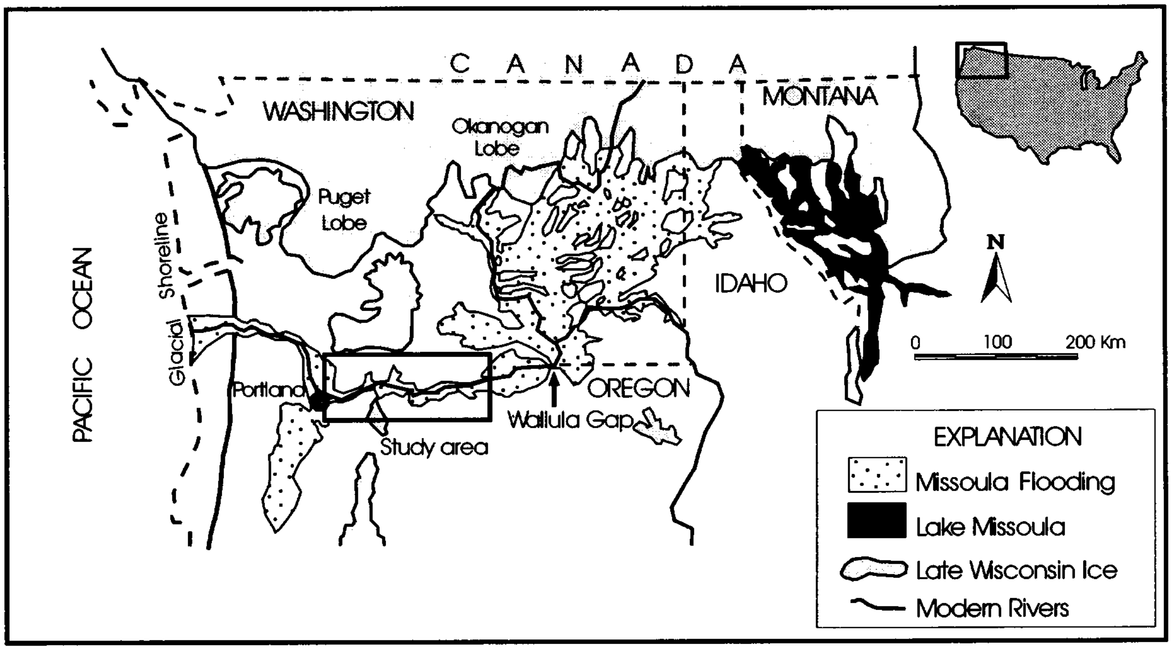


Figure 1. Location of the Columbia River Gorge.

mapped and measured patterns of erosion and deposition can, at local scale, provide a quantitative link between hydraulic conditions and resultant flood features.

STUDY REACH AND MODELLING RESULTS

Outbursts from glacial Lake Missoula were conveyed by multiple flow pathways across the eastern part of Washington State. Downstream of Wallulla Gap these flow paths converge into a single pathway, providing the best possible scenario for calculating palaeodischarges and the local hydraulic variables. The study area, known as the Columbia River Gorge, comprises a 200 km reach from Arlington to Portland (Figure 1). The bedrock geology within the Columbia Gorge is composed of the Yakima basalt or Miocene basalt flows, up to 600 m in thickness, overlain by a heterogeneous series of unconsolidated sandstone and conglomerate known as the Dalles Formation (Piper, 1932). Geologic indicators of the largest flood indicate flow widths varying from 12 km at the Dalles and Hood River expansion to 2·5 km at the constrictions through the Cascaces. During the largest flood, maximum flood stages of 310 m were recorded at the Dalles (maximum flow depth exceeding 275 m), and there is evidence for maximum stages exceeding 340 m (flow depth of 290 m) at the upstream end of the reach near Arlington. Across the Cascades, between Hood River and Portland, the flow was confined to a long and narrow canyon until its debouchment into the Willamette Basin. Here, the floods built a large expansion bar that Bretz (1925) named the ‘Portland delta’. In this reach, the flood maximum stages dropped from 285 m at Hood River (maximum flow depth exceeding 270 m) to 105 m a.s.l. at Portland (maximum flow depth exceeding 100 m). The intervening constriction in the apex of the Portland delta, near Crown Point, apparently regulated flow exiting the Gorge and significantly influenced the water-surface profile as far upstream as the Pasco Basin.

The step-backwater calculations, in conjunction with the geological evidence of maximum flow stages, indicate a peak discharge for the largest Missoula flood of $10\times10^6\,m^3\,s^{-1}$ (Figure 2A). The calculations assumed subcritical flow conditions except at three cross-sections located in the entrance of the flow into the Willamette Basin, where the water surface drops almost 100 m in 10 km. Upstream of Hood River, the calculated water-surface profile closely matched the geologic evidence of maximum stages, reflecting the high degree of

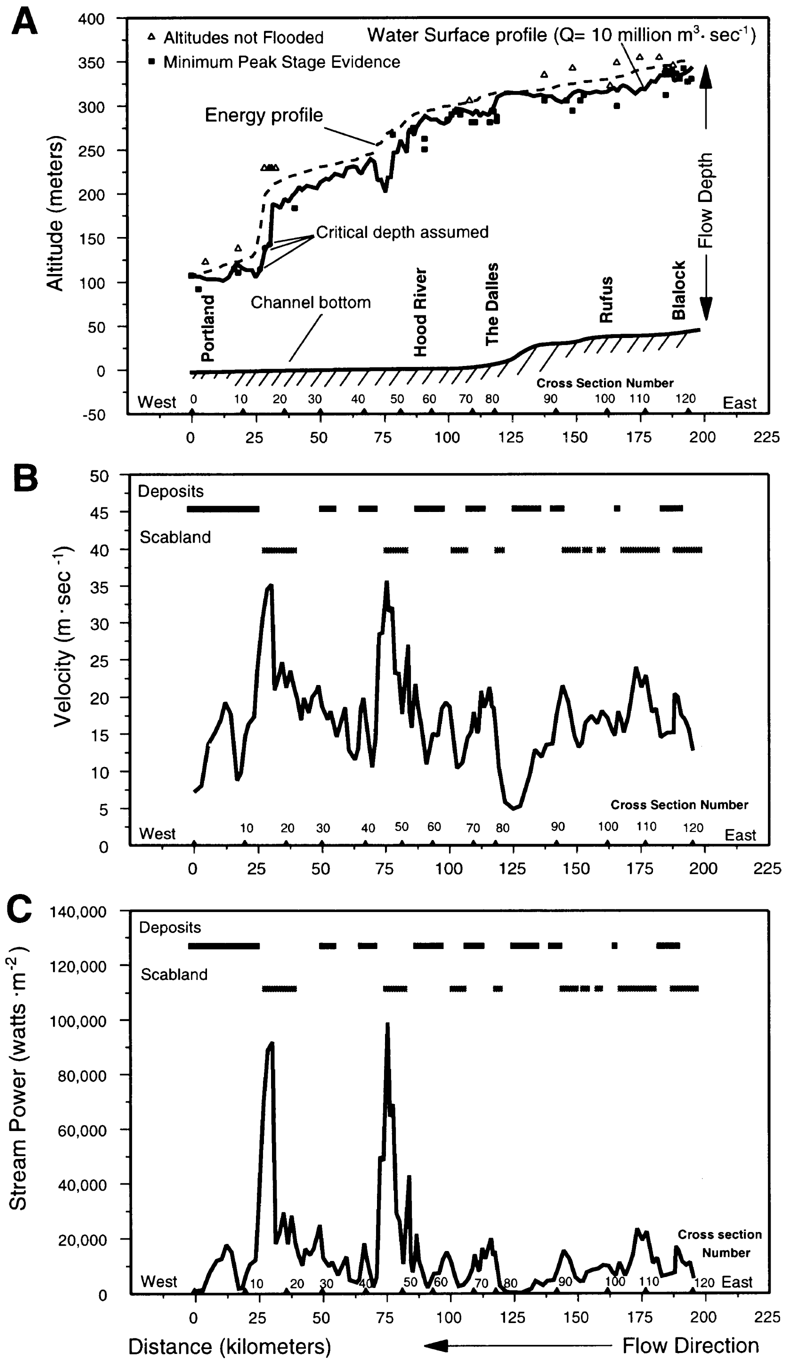


Figure 2. Down-channel variations of important hydraulic parameters as calculated from the step-backwater modelling. (A) Water-surface and energy-surface profiles. (B) Channel velocity and relationship to erosional and depositional features. (C) Maximum unit stream power values and relationship to erosional and depositional flood features.

precision in the modelling. Downstream of Hood River there is a lack of geologic evidence corroborating the modelling results. However, in the Willamette Basin in the geologic indicators are consistent with the calculated water-surface profile at peak discharges of $10\times10^6\,m^3\,s^{-1}$ (Figure 2A). The calculated peak discharge is equivalent to the discharge proposed by O'Connor and Baker (1992) at Wallula Gap, 110 km upstream of

Arlington. Other discharge estimates for the largest Missoula flood at Wallula Gap are $9\times10^6\,m^3\,s^{-1}$, calculated by Baker (1973a, pp. 17–22), and $12{\cdot}5\times10^6\,m^3\,s^{-1}$, obtained by Craig and Hanson (1985, pp. 52–53). For comparison, the maximum measured discharge for the Columbia River at the Dalles was $35\,000\,m^3\,s^{-1}$ during the 1894 flood.

FLOW ENERGY EXPENDITURE

The total energy expended by the flood releasing $2167\,km^3$ of water from an altitude of 1265 m (Pardee, 1942) to sea level was approximately $2{\cdot}5\times10^{19}$ J, equivalent to 35 per cent of the annual US energy production. The time duration of the flood was estimated to be approximately one week (Baker, 1973a). In terms of total energy expended, the Bonneville flood (O'Connor, 1993) was five times greater than the largest Missoula flood, although this energy was expended in approximately six weeks. The rate of energy expenditure during the largest Missoula flood was only exceeded by the Late Pleistocene superflooding of the Altay Mountains in Siberia (Baker *et al*., 1993) where 2×10^{19} J was expended within half the time duration of the Missoula's largest flood. Although only a small portion of the total flow energy would be applied to develop landforms, the rate of energy expenditure provides a useful index of the ability to perform geomorphic work.

In the step-backwater calculations the sum of a flow's potential and kinetic energy must equal that of a downstream cross-section less any energy losses between sections. The energy loss between cross-sections is calculated as the sum of frictional losses and form losses and is represented by the head loss (h_e):

$$h_e = L S_f + C(\alpha_2 v_2^2/2g - \alpha_1 v_1^2/2g)$$

where L is the distance between cross-sections, S_f is the local friction slope calculated as a function of Manning's n, C is the energy-loss coefficient relating to the channel expansions and contractions, v_1 and v_2 are flow velocities for the downstream and upstream cross-sections, α is the velocity-head coefficient accounting for non-uniform velocity distribution in a subdivided channel, and g is the gravitational acceleration.

In alluvial stream channels, where boundaries are easily eroded, the channel tends to establish a width, depth and gradient that minimizes the energy dissipation rate (Chang, 1979; Thorne *et al*., 1988; Simon, 1992). The concept of minimum energy dissipation rate states that the channel geometry changes to minimize non-uniformity in energy expenditure. In the long term, alluvial systems should attain an equilibrium between the driving forces (stream power) and the resisting forces (resisting power) close to the threshold of critical power (Bull, 1979). As in alluvial channels, hydraulic geometry in bedrock canyons probably represents an equilibrium between hydraulic regime and the geologic environment (Begin and Schumm, 1984). For instance, in narrow bedrock canyons the presence of undulating channel walls tends to minimize variance in energy expenditure downstream in a manner analogous to that commonly attributed to bedforms (Wohl, 1994). In the Columbia River Gorge, the lack of correspondence between channel geometry and estimated stream power is reflected in the non-uniformity of energy release along the study reach. The major gaps in energy dissipation were produced between Hood River and Portland, where the water surface profile dropped from 285 m at Hood River to 105 m at Portland over a distance of 80 km. In contrast, upstream of Hood River there is a constant rate of energy loss broken by minor gaps located within constriction reaches. Therefore, the response of the resisting power to changes in stream power of cataclysmic flooding is not continuous, and the threshold of critical power is overcome at discrete values of stream power.

EROSIONAL LANDFORMS

The erosional features developed in the Columbia Gorge are mainly controlled by the bedrock composition (strength) and the energy expended by the flow (stream power). The most spectacular erosional features are located on the Miocene basalt flows because of their dense planes of weakness with variable resistance to flood erosion (Baker, 1973b). Within a lava flow the cooling joints make the basalt readily erodible by the plucking action of the floodwater. Trimble (1950) and Baker (1973a) pointed out that the planes of weakness within the basalt bedrock, such as the cooling joints and basalt flow contacts, were an important influence on fluvial

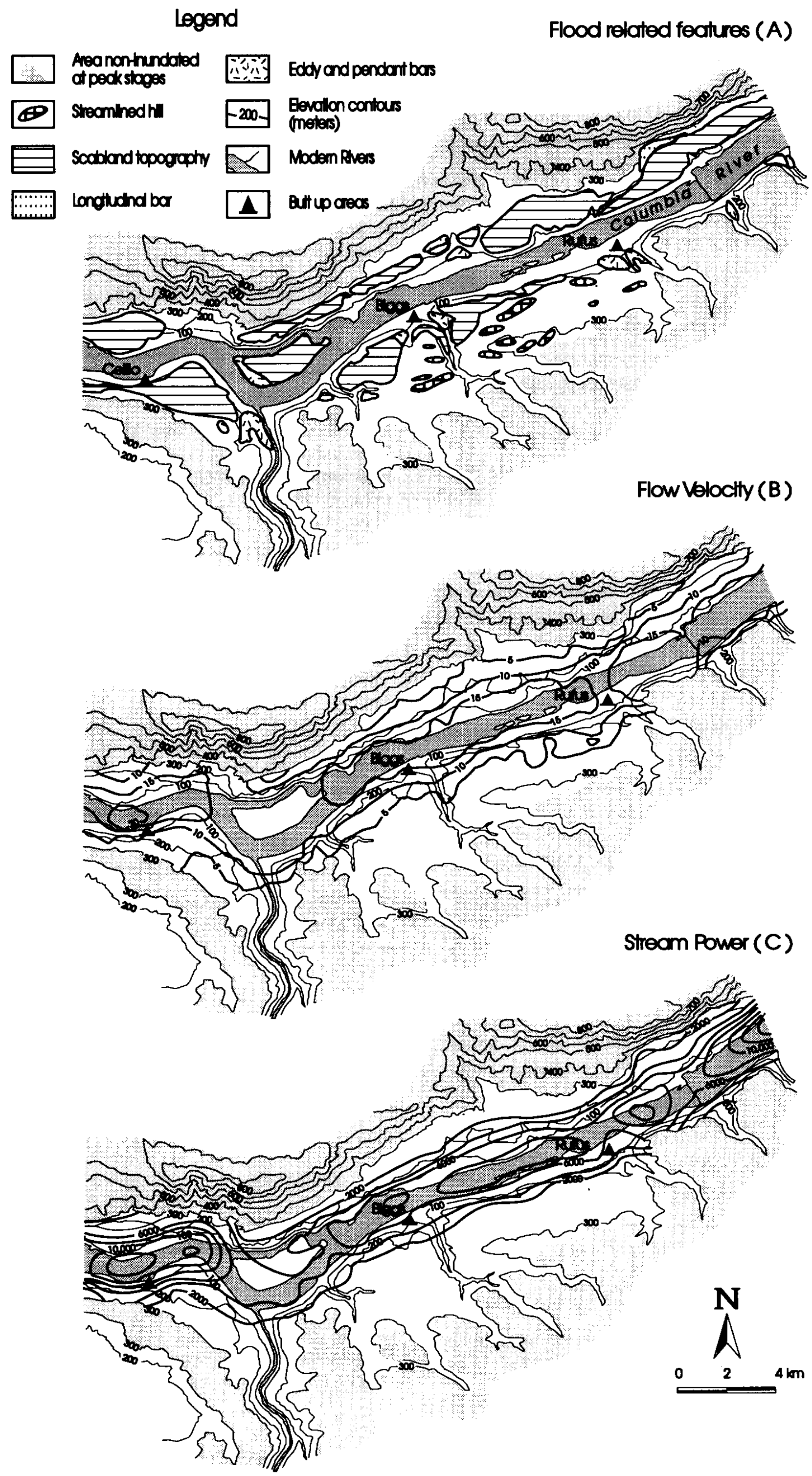
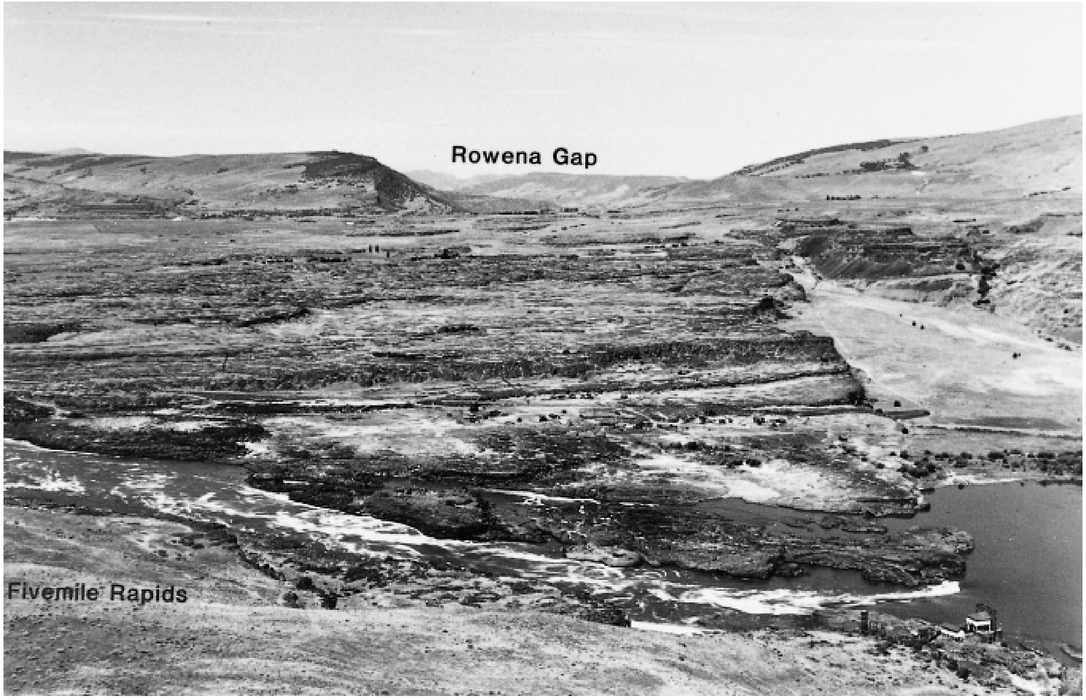


Figure 3. (A) Map of Missoula flood features, inundated area and topography for the Celilo–John Day River reach of the Columbia River Gorge. (B) Spatial distribution of flow velocity obtained from local hydraulic parameters calculated by the step-backwater modelling. (C) Stream power distribution obtained from local hydraulic parameters.

Figure 4. Rapids between Celilo Fall and the Dalles before inundation in1956 by the Dalles Dam. The topography of this reach within the Columbia River Basalt group consists of inner channels in the foregroung, with butte-and-basin ‘scabland’ topography behind, and stripped basalts in the background. The Celilo Falls, on the right-hand side of the picture, had a sheer drop of about 6m. USGSA photograph by A. M. Piper.

(a)



(b)

Figure 5. (a) Panorama of 'the Dalles of the Columbia' before the construction of the Dalles Dam. From the bottom to the top, the picture shows inner channels, butte-and-basin topography and stripped basalts. The Rowena Gap is one of the major constrictions of the Columbia River Gorge. To the left of the Gap, the tree line shows the stage of the largest Missoula flood during peak discharge. USGS photograph by A. M. Piper. (b) Giant bar on the channel wall left by the Missoula floods, 4 km northeast of Rufus. This bar is about 2 km in length and about 100 m thick. On the right-hand side, the aluminium factory and the Columbia River show the scale.

erosional forms produced by the Missoula flooding. The scabland erosion is ubiquitous in the Columbia Gorge, producing numerous cataracts (e.g. scabland southeast of Celilo) (Figures 3A and 4). The erosional features developed on the unconsolidated conglomerates of the Dalles Formation are not as well preserved as on the basalt flows. The development of Holocene soils and the detrital nature of these deposits have allowed a faster modification of the erosional landforms. The most characteristic landforms produced on the Dalles conglomerates are streamlined hills occurring between the Deschutes River and John Day River (Figure 3A).

For the largest Missoula flood, the estimated average unit stream power at a cross-section varied from less than $1000\,W\,m^{-2}$ at Portland and the Dalles Basin, to nearly $100\,000\,W\,m^{-2}$ downstream of Hood River (Figure 2C). For comparison, similar calculations for historic floods in bedrock rivers range up to approximately $6256\,W\,m^{-2}$ in the Herbert Gorge, Australia (Wohl, 1992), and $12\,800\,W\,m^{-2}$ in the Narmada River, India (Rajaguru *et al*., 1995). The estimated unit stream power for the largest floods on the alluvial Amazon and Mississippi Rivers are approximately $12\,W\,m^{-2}$ (Baker and Costa, 1987). The two largest peaks of stream power and velocity are associated with the constrictions located at Benson State Park, near the debouchment into the Willamette Basin, and at the junction with the Little Salmon River, just downstream from Hood River (Figures 2B and C). Upstream from Hood River the maximum values of stream power range between 20 000 and $30\,000\,W\,m^{-2}$. These values are obtained at constrictions located east of the junction with the John Day River, at Celilo (Figure 4), at the Dalles (Rowena Gap; Figure 5A) and at Moiser. Average channel velocities above $15\,m\,s^{-1}$ and stream power above $15\,000\,W\,m^{-2}$ will initiate the development of scabland erosion at a cross-section (Figures 2B and C). These average values may not reflect the conditions associated with the flood features through the complete cross-section. Therefore, flow separation within a cross-section is needed for computing the local hydraulic variables. To a large extent, these local hydraulic variations control the nature and patterns of erosional features left by the flood.

EROSIONAL SEQUENCE

As pointed out by Baker (1973b) and Baker and Komar (1987), the erosional forms occur in an evolutionary sequence that is related both to the flood flow hydrodynamics and to the resistant characteristics of the bedrock. The scenario used by those authors to describe this qualitative model was the Columbia Plateau where the Miocene basalt flows are capped by loess hills. The five-phase model is initiated by exposing the underlying basalt and leaving occasional remnants of streamlined hills. Increasing the flow depth caused groove development by turbulent floodwater, potholes and subsequently butte-and-basin topography by vertical vortices or ‘kolks’. Eventually, headward migration of structural steps in the basalt developed prominent inner channels. In the Gorge, the scenario is slightly different since the Dalles Formation capped the basalt flows and the flood was channelled into a single pathway. However, a similar temporal sequence of erosional landforms can be described at different positions in the Columbia River Gorge. Streamlined hills were developed on high plains occupied by shallow flows during the flood. In deeper areas, stripped basalts, butte-and-basin scabland topography and inner channels were developed. Local hydraulics associated with these areas allowed preliminary quantification of physical parameters of the erosional sequence.

The relationships between local fluid dynamics and measured patterns of erosion and deposition can be inferred in reaches well characterized by the step-backwater modelling. The reach between Celilo and John Day River (Figure 3) was chosen because of its variety of erosional and depositional landforms, and because of the quality of the modelling indicated by the good agreement between the geological indicators and the calculated water-surface profile (Figure 2A). In this reach, the lower energy expenditure rates in the erosional sequence are represented by streamlined hills (Figures 3A and C) developed on unconsolidated deposits of the Dalles Formation. These streamlined hills are interchannel elements produced by a network of anastomosing channels with split and converged flow analogous to braided gravel streams. Streamlining has the effect of reducing the drag or resistance to a flowing fluid (Baker and Kochel, 1978; Komar, 1984). Between Biggs and Rufus (Figure 3A), the 305 m peak discharge stage of the largest Missoula flood overflowed the south side of the Gorge at 270 m altitude, eroding an extension of $60\,km^2$ of high plains. The high-water mark reconstruction indicates that the streamlined hills were eroded both fluvially and subfluvially. In the areas where streamlined hills have developed, the step-backwater calculations show flow depths of 0–40 m and velocities lower than $5\,m\,s^{-1}$

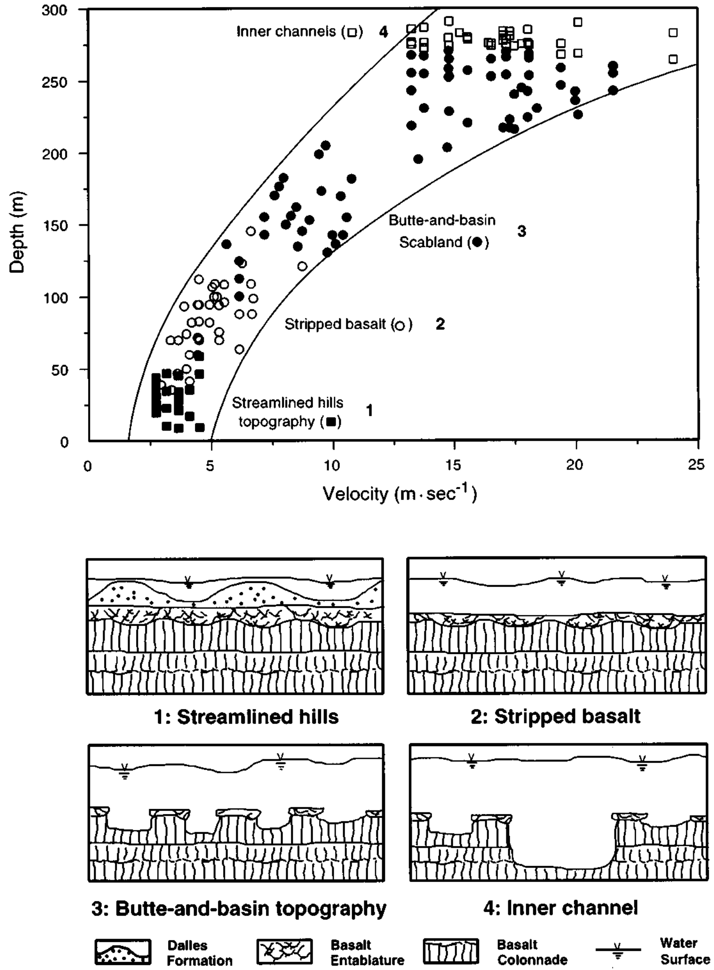


Figure 6. Relationships of local flow velocity and flow depth to erosional flood features. The numbers are assigned to steps of the hypothetical sequence of erosion for the development of scabland topography proposed by Baker and Komar (1987).

(Figures 3B and 6). Furthermore, the calculations indicate shear stresses below 100 N m^{-2} and stream powers lower than 500 W m^{-2}. The length/width ratio of the streamlined landforms has an average value of 3·2, which falls within the range of 3–4 presented by Komar (1984) in the minimization of drag forces. Similar lemniscate-like forms were described in the channelled scabland by Bretz (1923) and Baker and Nummendal (1978) as a residual form of loess 'islands' standing on the Miocene basalt. In the Gorge, these forms cannot be considered residual since the surrounding anastomosed channels were also scoured into the Dalles Formation.

With deeper and higher flow velocities the unconsolidated materials of the Dalles Formation were stripped away, uncovering the underlying Columbia River Basalt Group. The erosional features on bare basalt surfaces or 'scabland' (Bretz, 1923) occurred in a suite of morphologies that represent different stages of energy expenditure by macroturbulent flood flows (Figures 4 and 5A). At the low end of energy expenditure the resulting landforms correspond to intact bare basalt surfaces with scarce longitudinal grooves (Figure 6). These stripped basalts are mostly limited to ledges over 180 m in altitude located east of Rock Creek and between the Deschutes River and Rufus at the north side of the Gorge (Figure 3A). The step-backwater modelling shows that stripped basalt areas are associated with flow velocities of 3–9 m s^{-1} and flow depths between 25 and 125 m (Figures 3B and 6). Shear stress values range between 50 and 380 N m^{-2}, with an average value of 150 N m^{-2}. The calculated rate of energy expenditure or stream power per unit area ranges between 350 and 2800 W m^{-2}, with an average of 780 W m^{-2}.

Indicative of more intense erosion is the butte-and-basin scabland topography. This topography of mesas of basalt surrounded by closed depressions has resulted from plucking erosion by vertical flow vortices or kolks and by cavitation phenomena (Bretz *et al*., 1956; Baker, 1973a; Baker and Nummedal, 1978). Butte-and-basin scabland is ubiquitous along the Columbia Gorge below 180 m altitude, occurring on either basalt ledges, such as between Rufus and Rock Creek, or in basalt hillslopes of the Dalles and Lyle expansions (Figure 5A). Nevertheless, the most bizarre example is located west of the Deschutes River, at Celilo, where three basalt-flow flight surfaces, with an area of 4 km^2, were eroded by the floods, producing longitudinal grooves and numerous potholes (Figure 4). Although both processes were controlled by local flow lines and velocity perturbations, these landforms are associated with velocities ranging between 6 and 24 $m\,s^{-1}$, with an average value of 14 $m\,s^{-1}$, and depths varying between 100 and 270 m (Figures 3B and 6). The calculated shear stress associated with the basin-and-butte topography ranges between 250 and 800 $N\,m^{-2}$ and the stream power between 2000 and 20 000 $W\,m^{-2}$, with an average of 6500 $W\,m^{-2}$ (Figure 3C).

The basalt landforms related to the largest amount of energy expenditure are the inner channels. In the Columbia River Gorge, Bretz (1924) described multiple and anastomosed channels flowing in vertically walled basalt bedrock as the ‘Dalles type’ of channel. Indeed, before the closure of the Dalles Dam in 1956, the falls, pools and rapids between Celilo Falls and the Dalles were known by the name ‘the Dalles of the Columbia’ (Figures 4 and 5A). The inner channels were produced by headwater recession of scabland cataracts by plucking erosion and powerful kolks at plunge pool locations (Bretz *et al*., 1956; Baker and Nummedal, 1978). These falls reflect headwater retreat accomplished during the inner-channel formation. Intensive inner-channel retreat during the flooding produced a meander cut-off at Miller Island (Figure 3A), and a similar phenomenon, but at an incipient stage, was started at the Big Bend of the Dalles. The local energy expenditure by kolks and cavitation is difficult to estimate. The hydraulic calculations show that the minimum energy conditions that initiate processes producing inner channels were located at areas with water depths exceeding 250 m and flow velocities over 13 $m\,s^{-1}$ (Figures 3B and 6). The associated boundary shear stresses range from 300 to 1000 $N\,m^{-2}$ and the stream power per unit area varies from 4500 to 25 000 $W\,m^{-2}$, with an average value of 9000 $W\,m^{-2}$ (Figure 3C).

DEPOSITIONAL FEATURES

The scattered location of flood deposits along the Gorge reflects the limited sediment available in the Columbia Plateau versus the potential conveyance capacity of the flow. The depositional areas were controlled by local hydraulic conditions such as flow separation and decreasing stream power at the main channel. The flood deposits can be separated based on mode of sediment transport in tractive deposits and suspended-load deposits. Longitudinal bars, expansion bars and pendant bars were formed with ‘bedload’ materials, whereas both eddy and slack-water deposits were conveyed by the suspended-load fraction.

As indicated above, the step-backwater modelling implies some limitations; for example, flow is considered to be strictly in the downstream direction, providing only a partial characterization of the actual flow conditions. Therefore, depositional landforms located in areas of flow separation (eddy bars) and in the lee of resistant protrusions (pendant bars) are not physically characterized by the modelling. Only longitudinal bars and expansion bars can be accounted for by the analysis of local fluid dynamics. These bedforms were deposited within a main tread of the flow either in the Gorge or within channels that overflowed high divides. The longitudinal bars are located both downstream of constrictions, where a decrease in stream power included the deposition of materials tractively transported (e.g. downstream of Lyle or John Day River constrictions), and downstream of divide crossings, where the flow separation locally decreased the flow competence (downstream of Fairbanks Gap). Longitudinal bars are narrow, about 500 m, and elongate, with axial lengths from 2 to 3 km (Figure 5B). The larger axes are parallel to the main channel and attached to the canyon side. Altitude differences between the base and the crest range from 50 to 80 m. In contrast with longitudinal bars described by O’Connor (1993) in the Bonneville flood, the marginal channels between the bar crest and the flow margin are not well developed in the Columbia Gorge. Bars located below 180 m in altitude are very likely to be reworked either by multiple flood events or by different stages of a single flood.

FLOW VARIABLES ASSOCIATED WITH LONGITUDINAL BARS

As indicated by Church (1978) and O'Connor (1993), a flood deposit is the physical record of deposition in contrast to the physical record of initial motion, as has been the view of some researchers (Williams, 1983). Therefore, they represent depositional conditions or the limit of sustained transport. The assumption used for the Bonneville flood boulders by O'Connor (1993) states that a particle in transport will continue to move until the transporting ability (e.g. local values of velocity, shear stress, or stream power) of the flow diminishes (temporally and spatially) below a threshold necessary to maintain particle movement. Therefore, bar locations should indicate areas of decreasing transport capability (stream power) below the threshold maintaining particle movement. Although this concept seems to be theoretically consistent, the assumption that the sediments were emplaced during peak flow conditions is undoubtedly incorrect. The Missoula flood deposits could also have been emplaced either during the waning stages of the flood or by multiple Missoula floods (Waitt, 1980, 1984), most of these having low magnitudes (Baker and Bunker, 1985; Benito and O'Connor, 1995). The results of the step-backwater modelling at peak discharge may overestimate the energy conditions associated with these flood deposits. Analysis of the hydraulic variables related to the bar location should be considered as an approximation to the maximum values of flow energy associated with the depositional areas. Furthermore, for longitudinal bars to be related to the largest floods hydraulics, they should be located at high altitudes where minimum modifications are expected by smaller floods.

At peak discharge, calculated channel flow velocity associated with depositional areas was below $15\,\mathrm{m\,s^{-1}}$ and the estimated stream power was below $15\,000\,\mathrm{W\,m^{-2}}$ (Figures 2B and C). Since the depositional areas are located at the side of the main channel, local flow conditions reflect in a more accurate way the flow strength associated with flood deposits. Local flow hydrodynamic variables were analysed between Celilo and John Day River (Figure 3). At this reach, two longitudinal bars located above 190 m in altitude were emplaced by major floods. Three kilometres north of Biggs, the longitudinal bar is composed of at least three units of gravels and pebbles. The flood deposits are 70 m below the calculated water-surface elevation and were emplaced by a flood discharge of at least $6{\cdot}6\times10^{6}\,\mathrm{m^3\,s^{-1}}$. The longitudinal bar located 3 km north of Rufus is also composed of gravels and pebbles (Figure 5B). These deposits are 50 m below the calculated water-surface elevation and required at least $7\times10^{6}\,\mathrm{m^3\,s^{-1}}$ to be emplaced. Note that both bars were deposited by major floods of over $6{\cdot}5\times10^{6}\,\mathrm{m^3\,s^{-1}}$ and they possibly correspond to the largest flood, estimated at $10\times10^{6}\,\mathrm{m^3\,s^{-1}}$. At peak discharge, the calculated velocity associated with those longitudinal bars was less than $8\,\mathrm{m\,s^{-1}}$, shear stresses less than $200\,\mathrm{N\,m^{-2}}$, and stream powers were lower than $1500\,\mathrm{W\,m^{-2}}$ (Figures 3B and C).

The systematic evaluation of the depositional hydraulic conditions from the coarse particles left by the flood was not accomplished. However, the particle size distribution found at these bars compares with the flow competence relationships found by Baker and Ritter (1971), where a shear stress of $100\,\mathrm{N\,m^{-2}}$ is associated with an intermediate particle axis of below 20 cm.

DISCUSSION

Erosional processes developing actual bedrock channels are not well understood because of the inability to observe and measure the processes and rates, or to monitor long-term changes in their position. Indeed, there are problems in determining the temporal scales of erosional processes either over extremely long periods at very slow rates, or over extremely short periods at very high rates. As a consequence, there is little quantitative information regarding erosional thresholds for natural bedrock fluvial systems. The erosional landforms developed by the Missoula floods were produced over short time spans and in reaches of high rates of energy loss. The erosional sequence or geomorphic work accomplished by Missoula floods, described by Baker and Komar (1987), was developed by increasing either spatially or temporally the efficiency of power expenditure. The relationships between erosional features along the Columbia Gorge and the local hydraulic conditions calculated from the step-backwater modelling provide a useful tool in understanding these erosional and depositional landforms.

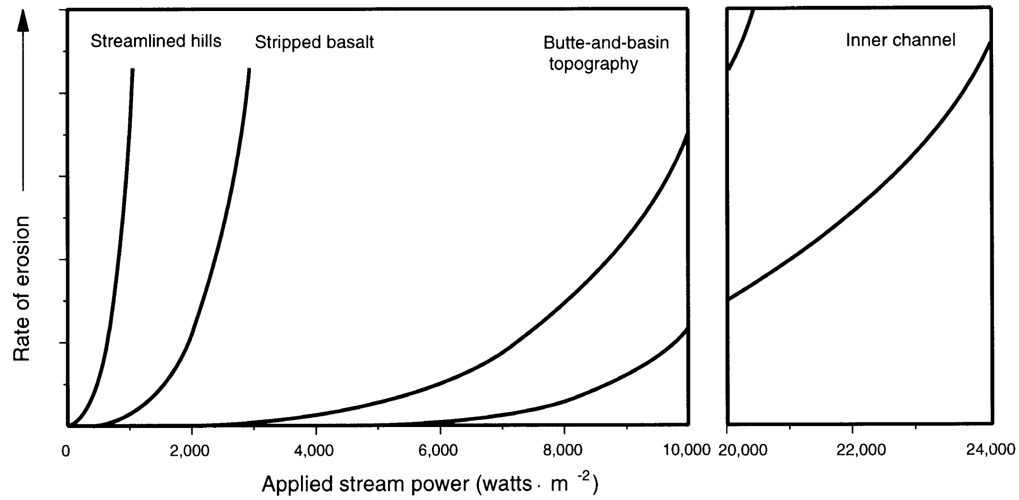


Figure 7. Generalized relationships between rate of erosion and applied stream power. Stream power thresholds for the development of erosional landforms are indicated. The effective stream power is the stream power exceeding these thresholds.

In order to initiate erosion, some threshold must be exceeded by the applied stress, below which no work is expended to develop erosional landforms (Figure 7). Therefore, the erosional sequence is not a continuous evolution, depending on total energy expenditure applied to the channel, but a discontinuous succession of landforms initiated at discrete rates of energy expenditure or thresholds of stream power. Bull (1979) defined the threshold of critical power in streams as a balance between opposing tendencies or the ratio between driving forces (stream power) and resisting power. The concept of threshold of critical power can be applied to the erosional sequence produced during Missoula flooding, with some modifications because in this case the ratio does not represent a balance between degradation and aggradation. When the stream power is less or equal to the resisting power, a situation of 'equilibrium' or non-erosion occurs, therefore, only for ratios larger than 1 will degradation conditions result. The stream power can be calculated from the step-backwater modelling and reflects indirectly the result of different erosional mechanisms such as longitudinal roller vortices, kolks and cavitation producing separation of the basalt columns. Although Baker and Komer (1978, p. 427) indicated that 'it is not yet possible to express resisting power in the same physical terms as stream power', we may estimate the resisting power in terms of critical power required to produce specific erosional landforms. Therefore, the estimated minimum stream power required for the formation of specific erosional landforms reflects the sets of variables acting in favour of the resisting power.

As indicated above, the critical stream power required to initiate erosion varied for the different landforms of the erosional sequence (Figure 7). The shear stress and stream power thresholds were estimated at 50 N m^{-2} and 350 W m^{-2} for stripped basalts, 250 N m^{-2} and 2000 W m^{-2} for basin-and-butte scabland topography, and 300 N m^{-2} and 4500 W m^{-2} for inner-channel formation. Streamlined hills may be developed in very low energy conditions associated with shear stress and stream power of over 100 N m^{-2} and 500 W m^{-2}, respectively. Erosion is possible only during catastrophic floods exceeding those thresholds of stream power below which no work is expended in erosion. Therefore, although multiple outbursts of Lake Missoula did occur during the late Pleistocene (Waitt, 1980, 1984, 1985; Baker and Bunker, 1985; Benito and O'Connor, 1995), only a few of these events had the required magnitude to overcome the threshold conditions and accomplish significant geomorphic work (Figure 8).

The concept of geomorphic effectiveness, as put forward by Wolman and Miller (1960), involves the frequency of occurrence as well as the magnitude of individual events. This concept was used to demonstrate that relatively more work is accomplished in modifying landscapes by frequent geomorphic events of low magnitude than by rare catastrophic events. Later, Baker (1977) demonstrated that landscape modification in some areas is only accomplished by large floods exceeding specific thresholds, and therefore the geomorphic

processes are effective only in the largest events. This is true in the case of the erosional landforms developed by the Missoula flooding, where landform changes were initiated only when critical values of stream power were exceeded (Figures 6 and 7). Similarly, if it is assumed that the rate of sediment transport is a power function of the causative fluid stress or power (Bagnold, 1977; Leopold *et al.*, 1964), then we may assume that once the energy expenditure needed to separate and entrain the particles from the jointed lava flows is achieved, the rate of erosion (q_e) is a function of the stream power:

$$q_e = k(\omega - \omega_c)^n$$

where k is a constant related to the characteristics of the material eroded, ω is the applied stream power per unit area, and ω_c is the critical stream power required to initiate the bedrock erosion. Similarly to the semi-arid drainage explained by Baker (1977), the effectiveness of the Lake Missoula outbursts is not measurable by its frequency but by its magnitude and rate of energy expenditure. Therefore, to initiate geomorphic work, some threshold of resisting power should be exceeded, below which the effectiveness is null. The difference $\omega - \omega_c$ can be defined as the effective stream power or rate of energy expenditure capable of performing geomorphic work and producing specific landforms.

Although the total energy expended by the tens of Lake Missoula outbursts was very high, effective geomorphic work could have been accomplished in only a few exceptionally large floods capable of exceeding the erosional threshold values (Figures 7 and 8). These critical values explain why the geomorphic evidence for the largest floods, such as basin-and-butte topography and inner channels, has not been erased by multiple smaller floods. In contrast, the depositional landforms left by the largest floods in the main channel at altitudes lower than 180 m have been reworked by the smaller floods. The threshold of stream power needed to transport sediment could be exceeded even during small floods. Only longitudinal bars occurring at high altitudes can be related to the hydrodynamic conditions of the largest floods.

ACKNOWLEDGEMENTS

The research for this paper was carried out during a postdoctoral fellowship, funded by the Spanish Ministry of Education, in the Arizona Laboratory for Paleohydrological and Hydroclimatological Analysis (ALPHA). The assistance, teaching and discussion of ALPHA colleagues have been crucial for its progress. I am indebted for

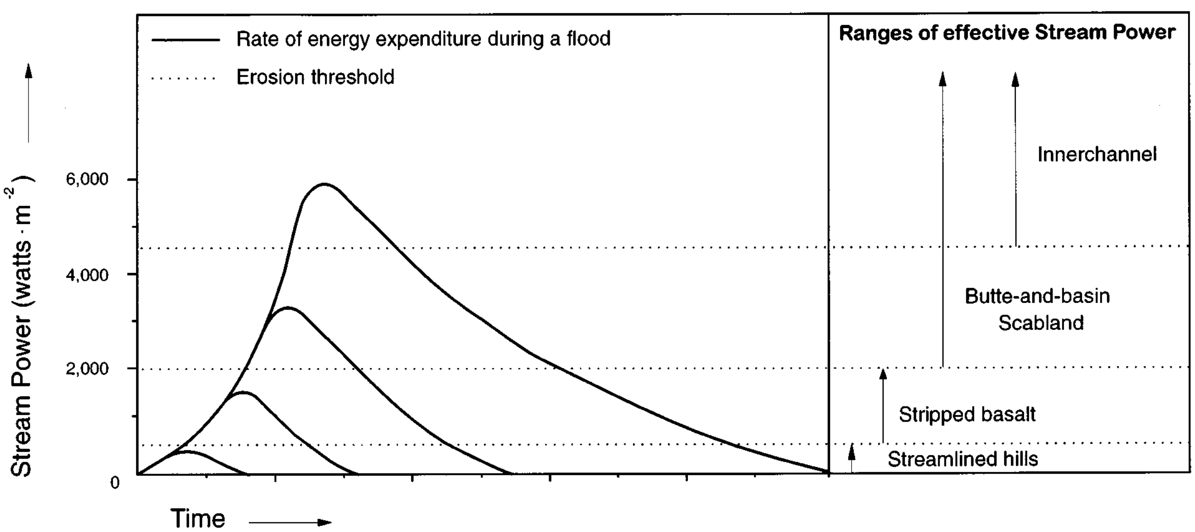


Figure 8. Stream power distribution within a cross-section during different flood hydrographs. Note that only a few floods are sufficiently large to exceed specific thresholds in the erosional sequence.

the inspiration, encouragement and support of J. O'Connor and V. R. Baker. I am grateful to the US Geological Survey, Cascades Volcano Observatory (Vancouver, WA), and particularly to J. E. Costa for providing logistical support during the field work. This project was improved by field discussions with R. B. Waitt of the US Geological Survey, regarding the sedimentological evidence for multiple Missoula Lake outbursts. The constructive reviews by Christopher Fitzjohn (University of Plymouth) and anonymous reviewers substantially improved the manuscript.

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